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Soil erosion in relation to climate change and vegetation cover over the past 2000 years as inferred from the Tianchi Lake in the Chinese Loess Plateau

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Graphical abstract

Highlights:

Six major periods with more than 20 serious soil erosion events are recorded in the Chinese Loess Plateau.

Regional high-intensity rainfall or flood events are the main cause of soil erosion.

Light vegetation cover aggravated soil erosion, whereas heavy cover reduced soil erosion.

Vegetation restoration in the catchment is extremely critical to control soil erosion in the Chinese Loess Plateau.

Abstract

Understanding the response of past soil erosion to climate change and human disturbances is quite essential for managing and preventing soil erosion in the future. Here, we report a high-resolution soil erosion sequence during the past 2000 years

based on grain-size data from a sediment core GS07A in Tianchi Lake in the Chinese Loess Plateau. These grain-size data are analysed by end-member modelling and further deciphered by detailed modern investigations of surface sediments, surface soils, and modern dust samples aided by published grain-size distribution characteristics of different depositional provenances and processes. Six strong soil erosion periods, including more than 20 serious soil erosion events, are inferred at 150-300 AD, 450-750 AD, 900-1100 AD, 1200-1400 AD, 1650-1800 AD, and 1950-2000 AD. These soil erosion records correspond well with intensive flood events in downstream rivers and the neighbouring Longxi area documented from the historical literature, suggesting that regional high-intensity rainfall or flood events are the main cause of soil erosion. Vegetation cover in the catchment also influenced the soil erosion: light vegetation cover aggravated soil erosion even in the periods with a low frequency of flooding, while heavy cover reduced soil erosion even in the periods with a high frequency of flooding. Our results highlight the role of sustainable land use and vegetation restoration around the catchment in controlling soil erosion in the Chinese Loess Plateau.

Keywords: Grain-size; Soil erosion; Vegetation cover; Loess Plateau; Tianchi Lake;

1. Introduction

Soil is an important component of the Earth System. It controls the hydrological, biological and geochemical cycles, and also provides ecosystem services for human

(UNDESA, 2013; Keesstra et al., 2016; Solomum et al., 2018). Erosion of topsoil is a widespread and major threat to terrestrial ecosystems (Singer and Warkentin, 1996; Belyaev et al., 2005). Soil erosion can lead to the severe decline of land quality and agricultural productivity, and even affect deterioration of the ecological environment (MEA, 2005; Tsunekawa et al., 2014; Keesstra et al., 2016). Thus, a better understanding in mechanism of soil erosion would help to reduce the loss of soils, and support socio-economic and ecological sustainable development under the expected climate warming (IPCC, 2013; UNDESA, 2013; Keesstra et al., 2018a; Halbac-Cotoara-Zamfir et al., 2019). Numerous studies have shown soil erosion is not only affected by natural factors such as soil type, gradient, surface river, and heavy rainfall (Keesstra et al., 2009; Parris et al., 2010; Keesstra et al., 2014; Jia, 1990; Xin et al., 2011; Sun et al., 2013), but also influenced by human activities such as farming, land abandonment, deforestation and introduction of new pastrue species (Gyssels and Poesen, 2003; Fattet et al., 2011; Fu et al., 2015; Antoneli et al., 2018; Cerdàet al., 2018). Because of the connectivity between land use, vegetation cover and soil erosion, anthropogenic activities can severely aggravated soil erosion through the changes in connectivity (Keesstra et al., 2018b). However, most of theses studies have been focusing on modern soil erosion monitoring and modelling, less attention is paid to the past soil erosion history, especially for centennial- or millennial timescales, as the past is the key to understanding the present and predicting the future (Smol, 1992; Dearing et al., 2008).

The Chinese Loess Plateau is located in the transition zone between semi-humid

and semi-arid, and highly susceptible to erision. It has been suffered from severe soil erosion for a long period with more than 80% of its area now (Yang and Yu, 1992; Wang et al., 2006; Wei et al., 2006; Tsunekawa et al., 2014; Feng et al., 2017). At centennial- or millennial timescales, researches on past soil erosion in response to climate change and human activities are very limited (He et al., 2004, 2006; Liu et al., 2018). Lacustrine sediment, as a collector of past climatic and environmental changes, can not only record long-term, continuous, high-resolution and undisturbed information of regional rainfall, vegetation cover and human disturbances (Zhao et al., 2010; Xiao et al., 2015; Chen et al., 2015), but also provide information about soil erosion processes (Oldfield et al., 1990; Dearing et al., 2008). Thus it owns great advantage in reconstructing soil erosion history and understanding its driving mechanism (Smol, 1992; Dearing et al., 2008; Dearing, 2013). Especially for alpine lake, the catchment characterized by steep and bare slopes makes the surface soils more susceptible erosion, which can be better recoded in alpine lake sediments (Giguet-Covex et al., 2011; Arnaud et al., 2012; Yu et al., 2016). Tianchi Lake is located near the top of Liupan Mountain which is the geographic boundary between Longdong and Longxi sections of the Chinese Loess Plateau (Liu, 1985). Previous studies showed Tianchi Lake have a stable, continuous, high-resolution and reliable chronological sediment cores with varve sequence, and the detailed records of past climate changes and vegetation evolution were reconstructed based on a series of proxies such as TOC, elements and pollen records (Zhou et al., 2010; Zhang et al., 2010; Zhao et al., 2010). Besides, historical literature and cereal-type pollen also

indicate the lake catchment has a long history of strong human activities for the last few thousand years (Bettinger et al., 2010; Zhang et al., 2010). These evidence indicates that Tianchi Lake could provide ideal materials to understand the response processes of soil erosion to climate changes, vegetation cover and land use at centennial- or millennial timescales. In this paper, aided by published regional climate change and vegetation cover data, we aim to investigate the connection between soil erosion and climatic/environmental changes based on grain-size distribution through the end-member modelling analysis for the past two millennia.

2. Study area

Tianchi Lake (lat.35°15′53″N, long.106°18′33″E, elevation 2430 m a.s.l.) is located near the top of Liupan Mountain in the south-west Chinese Loess Plateau, ~30 km east of Zhuanglang country, Gansu province (Fig.1a). The lake has a surface area of 0.02 km^2 with a maximum water depth of ~8 m. It is a freshwater lake and fed by precipitation and groundwater with a seasonal surface outflow on the west side. Instrumental data show the annual mean temperature is ~8.2 °C, the mean annual precipitation is ~495 mm with 87% of precipitation falling in the summer monsoon season from May to October based on 1976-2005 data from nearby Zhuanglang Meteorology Station (at 1615 m a.s.l.) (Fig. 1b). Also, Liupan Mountain area is the source region of Qingshui River, Hulu River, and Jinghe River, the tributary of Wei River.

Liupan Mountain is located near the current boundary between semi-arid and

semi-humid climate, and influenced by the East Asian summer monsoon, the Indian summer monsoon, and the westerlies. The catchment of Tianchi Lake is situated in the western of the Liupan Mountain and there are several small gullies leading the inflow into the lake (Fig. 1c). The soils in the lake catchment are dominated by gray-cinnamon soil which was formed by weathered residuals and slope deposits of sandy mudstone, shale and limestone (Zhang, 2016). The gray-cinnamon soil consists of litter layer, soft humus layer, mineral organic layer and deposit layer from top to bottom. This soil is very thin with quite fine particle, and these feature would make soils extremely susceptible to erosion (Wang et al., 2006; Tsunekawa et al., 2014). The modern vegetation on the lake catchment is dominated by shrubs and steppe, mainly consisting of *Corylus*, *Picea*, *Spiraea*, *Hippophae, Berberis, Rosa, Melilotus*, *Dendranthema*, *Artemisia*, *Roegneria*, *Phragmites, Stipa,* and *Festuca* (Zhao et al., 2010).

3. Material and methods

3.1 Sampling

In 2007, an 11 m-long sediment core (GS07A) was retrieved from the deepest part (8.2 m) of Tianchi Lake using a piston corer from a UWITEC platform (Zhou et al., 2010). The core was taken back to the laboratory in PVC pipes and stored in a refrigerator below 4 °C. The upper ~4 m of the core was chosen to reconstruct climate changes and soil erosion spanning the past 2000 years. Meanwhile, 5 lake

surface sediments, 21 surface soils within the lake catchment including 12 on the lakeshore and 9 on the hillside, and 2 dust samples from around Liupan Mountain were collected in August 2014 (Fig. 1c).

3.2 Chronology control

Seven terrestrial plant macrofossils from the top 4 m of core GS07A were selected for radiocarbon dating using accelerator mass spectrometry (AMS) at Beijing University (Table. 1). All the ^{14}C dates were calibrated to calendar ages using the IntCal13 calibration data set (Reimer et al., 2013) in the CALIB 5.0.1 program. An age-depth model was developed with smooth fitting using CLAM 2.2 (Blaauw, 2010) in the R software. The dating results are reported by Zhou et al. (2010) and Sun, (2011).

3.3 Grain-size analysis and end-member modelling

In the laboratory, the top 4 m of core GS07A was sliced at 1 cm intervals (~5-year resolution) for grain-size analysis after freeze-drying. Following the conventional chemical pre-treatment procedure (Konert and Vandenberghe, 1997; Sun et al., 2002), 400 subsamples were treated to remove organic matter with hydrogen peroxide (H_2O_2) and to remove carbonates with hydrochloric acid (HCl), and then dispersed with 10 ml sodium hexametaphosphate ($(NaPO_3)_6$) solution and ultrasound for 5 min. Grain-size distributions were measured using a Malvern Mastersizer 2000 laser grain-size analyser, with a size range from 0.02 to 2000 μm and 0.1φ interval resolution. Detrital contents were also calculated based on the

remaining weight after removing organic matter and carbonate content by combustion at 950 °C for 2 hours, then to estimate the contributions of terrestrial detrital materials from the catchment. All experiments were conducted in the Key Laboratory of Western China's Environmental Systems, Lanzhou University.

The grain-size distribution data are analysed by end-member modelling method to decomposed potential optimal end-members based on the following steps: First, grain-size data were averagely divided into 100 grain-size fractions in logarithmic intervals from 0.02 to 2000 μm. 400 subsamples and 100 grain-size fractions formed a matrix of 400×100 (m×n: measurements×variables). Second, this matrix dataset are input into the end-member modelling algorithm based on open-source code published by Dietze et al. (2012) running in MATLAB2016 software. The detailed description and end-member modelling script refer to Weltje et al. (1997, 2003) and Dietze et al. (2012). The potential optimal end-members are identified when the model stabilises with the least number of end-members and the highest explained cumulative variance. These criteria can be assessed through several statistical parameters such as mean explained variance (mean total r^2) for each end-member and explained variance for each grain-size and each sample, which can estimate the stability of the model and degree of fitting of terminal end-members.

3.4 Historical flood events

Liupan Mountain is the source region of Jinghe River, Qingshui River, Hulu River, and Wei River (Zhou et al., 2010). A series of flood events were investigated in downstream rivers, including Jinghe River, Weihe River, Fenhe River, the middle

reaches of the Yellow River, etc, from historical documentary records, and the flood frequency per 100 yr was further calculated (Table 2, Shi et al., 2002).

4. Results and interpretation

4.1 Lithology and chronology

The top 4 m of the sediment core mainly consists of fine-grained clay laminated sediments with several short intervals of obvious brownish coarse-grained silt or sand fractions. Based on the age model, the core has a mean sedimentary rate of 1.85 mm y^{-1}. The age at 5 cm in the GS07A core is consistent with the date in the GS14A core (red dot in Fig. 2) established by ^{210}Pb and ^{137}Cs dating based on grain-size data comparison, which was reported in Zhang (2016). The mean sedimentation rate for the entire core is 0.19 cm/yr, which gradually increased from 0.1 to 0.21 cm/yr from 0 AD to 1500 AD, and further increased to 0.33 cm/yr at 2000 AD (Fig. 2). All dates discussed in this paper are in calendar years.

4.2 The optimal end-members

To extract robust optimal end-members, the model was run in different numbers of end-members. The potential optimal number of end-members lies at the inflection point of end-members–mean total r^2 curve (Yeomans and Golder, 1982; Weltje et al., 1997; Prins et al., 1999; Dietze et al., 2012; IJmker et al., 2012). As shown in Fig. 3a, the mean total r^2 decreases after six end-members, suggesting the model tends to become unstable. Thus, the maximum number of end-members should be limited to six. The potential number of end-members (mean total r^2) is thus three (0.75), four

(0.80), five (0.82), or six (0.81). The mean total r^2 of two end-members (2EMs) and three end-members (3EMs) are <0.4 and <0.5 in most grain-size ranges, respectively (Fig. 3b). The goodness of fitting in 2EMs and 3EMs is extremely poor for <2 μm, 5-50 μm, and >100 μm, and unable to reach the requirement of fitting. Although five end-members (5EMs) and six end-members (6EMs) show a little better in several grain-size ranges, the mean total r^2 of both are still low for <2 μm and >100 μm fractions. For four end-members (4EMs), the mean total r^2 for each grain-size range improves greatly with explained variance more than 0.98 (Fig. 3b). The mean total r^2 of all grain-size ranges is 0.92, and thus all grain-size ranges are well represented by the 4EMs model (Fig. 3c). The mean total r^2 of all samples is 0.95 with about 94% of the samples higher than 0.9, indicating a majority of samples can be well explained by the 4EMs model (Fig. 3c, d). Therefore, the 4EMs model was selected to decompose grain-size data in our end-member analysis model.

4.3 Result of end-members

The grain-size distributions of final four end-members are shown in Fig. 4a. EM2, EM3, and EM4 have a unimodal distribution with dominant modes at 2 μm (fine clay), 30 μm (medium silt), and 10 μm (fine clay), respectively. EM1 has a multi-modal distribution with two prominent modes in the coarse sand (400 and 100 μm) and one minor mode in the fine clay (8 μm) fractions. Although all these end-members have one or two extremely minor modes, their contributions are quite low. Also, these four end-members well simulate the grain-size distributions in all sediment samples which shows three obvious peaks at <2 μm, 5-40 μm, and >100 μm

(Fig. 4a).

The loading scores of four end-members were also calculated to show the contributions of each end-member to each sample (Fig. 5). A majority of samples contain three or four end-members. For the whole core, EM1, EM2, EM3 and EM4 represent 2.3 %, 6.9 %, 26.9 %, and 63.9 % of end-member loading, respectively. EM4 dominates the sediment sequence with over 75% of the samples higher than 50%. EM3 contributes the second largest loading score with a large change ranges from 0 to 80%. EM2 accounts for quite small loading scores, and is completely absent in half of the samples. EM1 has the lowest average loading, but it contributes extremely high proportion during several short intervals. The mean grain-size does not show a significant change trend over the past 2000 year, but displays several obvious abrupt intervals (Fig. 5). Also, the proxy of coarse grain-size fraction (>100 μm) is calculated based on the grain-size distributions of surface soil sample to compare with the change in loading score of EM1 (Fig. 5). The distinct high peaks of mean grain-size and coarse fraction (>100 μm) show a quite good correlation with the high proportion of EM1.

5. Discussion

5.1 Interpretation of end-members

Grain-size distributions contain sedimentological information such as sedimentary sources and transport processes, and a certain typical grain-size

distribution can represent a specific depositional environment (An et al., 1991; Sun et al., 2002, 2003). Similarly, grain-size distributions of each end-member also provide the information about sediment source or transport processes (Parris et al., 2010; Dietze et al., 2012; Zhang et al., 2017). The final four end-members (EM1, EM2, EM3, EM4) can be deciphered to track the climatic and environmental changes based on specific grain-size distribution characteristics of different depositional provenances and processes, and aided with detailed modern investigations of surface sediments, surface soils, and dust samples (Sun et al., 2002; IJmker et al., 2012).

Among the four end members, EM4 has the highest loading in most samples with an average of 63.9%, showing a narrow mode in the fine silt fraction. Its peak grain-size and distribution pattern coincides well with the grain-size distribution of surface sediments near core GS07A (Fig. 4a, b). The similar grain-size distribution, dominated by a fine silt fraction with a mode at about 10 μm, can be found in the surface soils from lakeshores and hillsides, indicating EM4 may be mainly from surface soils. Moreover, the typical grain-size distribution of lacustrine sediments in slack water conditions also shows a narrow peak in the clay component with a peak of about 10 μm (Sun et al., 2002; An et al., 2012). Hence, EM4 could represent the sediment fraction of suspended fine particles deposited under slack water conditions from surface soil. EM3 contributes the second largest loading score, showing a narrow mode in the medium silt fraction with a peak at 20-40 μm. The grain-size distribution is different to that of surface soils, but similar to that of the local dust samples collected around Liupan Mountain area (Fig. 4c). A series of local dust

samples, collected at six different sites around the Chinese Loess Plateau in different seasons during 1995-1996 and 2000, also record the similar main mode centred at 20-40 μm (Sun et al. 2003). The good consistency between EM3 and local dust samples collected in the local Liupan Mountain and even larger Loess Plateau areas supports that local dust likely mainly contributes EM3 fraction. EM2 has the finest particle size at ~2 μm, which is similar to the grain-size distribution of the remote atmospheric dust (Fig. 4d). These dust fractions are mainly transported from the soured region to the downwind areas including the Loess Plateau by the westerly circulation at a very high level. Previous studies also suggest the grain-size distribution of remote atmospheric dust is dominated by a peak at 1-5 μm (Rea, 1994; Sun et al., 2003), consistent with the grain-size distribution of EM2 in the sediments. Thus, the fraction of EM2 may be mainly derived from the remote atmospheric dust.

Coarse sand EM1 has a multi-modal distribution with a fine mode at ~8 μm and a coarse modes at ~100-400 μm. This multi-modal distribution is similar to the grain-size distribution of surface sediments and soils, suggesting that the EM1 may derived from the surface soils in the catchment (Fig. 4a, b). The similar grain-size distribution between EM4 and fine fraction of EM1 also supports this. Although EM1 consists of two peaks, the coarse part should dominate the change of end-member loading scores due to the distinct negative relationship in the contribution between EM1 and EM4 (Fig. 5). Also, the coarse grain-size fraction (>100 μm), calculated based on the grain-size distribution of surface soils, shows good correlation with the high proportion of EM1. The increase of coarse particle fractions in the sediments may be

related to falling lake levels or enhanced inflow into the lake (Chen et al., 2004; Simonneau et al., 2013; Sheng et al., 2015; Xiao et al., 2015). Tianchi Lake is very small and fed only by precipitation and groundwater with a seasonal surface outflow on the west side, the lake level did not experience large fluctuations during the past two millennium (Zhou et al., 2010; Yu et al., 2016). The influence of lake levels in the variations of grain-size distribution is quite limited. By contrast, the significantly increased coarse particle materials may be attributed to high-energy inflow related to heavy rainfall, which is also supported by good correlation between increased coarse fraction and high peaks in mean grain-size. Since the mean grain-size can indicate the hydraulic energy of inflow for a very small lake with the stable water levels (Chen et al., 2004; Schillereff et al., 2014). The enhanced inflow caused by high-intensity rainfall and flood events, would significantly erode the detrital materials of surface soils and deposited into flood layers in the lake sediments. The depositional process of flood layer is characterised by a fining-upward grain-size pattern with coarse particles at the bottom grading up fine-grained laminate at the top (Prins et al., 2010; Gilli et al., 2013; Schillereff et al., 2014). At the beginning of a heavy rainfall or flood event, high-density inflow would carry a large amount of coarse particles from surface soils into the lake basin. Then, as the hydraulic energy of inflow gradually decreased, the suspended finer particles began to slowly settle down and eventually covered the coarse particles (Gilli et al., 2013). Similarly, the coarse part of EM1 was firstly eroded into the lake from the catchment by extremely strong inflow, and the fine part of EM1 was then slowly deposited when the inflow runoff was relatively

weak. Moreover, stronger soil erosion related to higher energy of inflow may erode more coarse detrital materials and result in higher loading score of EM1 in the lake sediments. Thus, the EM1 fraction was caused by high-energy inflow, and can record the depositional processes and intensity changes of flood events and/or soil erosion in the catchment.

5.2 Relationship between soil erosion and climate change/vegetation cover

More than 20 strong soil erosion events were identified in the sediment core through end-member modelling analysis of grain-size data during the past 2000 BP. These events could be grouped into six strong soil erosion periods at 150-300 AD, 450-750 AD, 900-1100 AD, 1200-1400 AD, 1650-1800 AD, and 1950-2000 AD (Fig. 6a). Detrital contents could indicate the relative contribution of terrestrial coarse-grained materials eroded from the lake catchment. Six periods of high detrital contents were also recorded at 150-300 AD, 500-750 AD, 900-1100 AD, 1200-1400 AD, 1650-1800 AD, and 1950-2000 AD, showing good correlation with the reconstructed soil erosion intervals, which supports soil erosion was mainly from the coarse-grained soils washed by strong inflow into the lake (Fig. 6a, b). There are many factors such as climate change, flood events, vegetation cover and human disturbances causing soil erosion (Zhang et al., 2009; Zhou et al., 2010; Gilli et al., 2013; Schillereff et al., 2014). Some studies suggest that soil erosion is more serious in humid periods than that in dry periods (Jia et al, 1990; Zhang et al., 2009; Xin et al., 2011), while others argue that dry climatic conditions or climatic transitions are more likely to cause severe soil erosion (Tang et al., 1991; Huang et al., 2007, 2011).

Published paleoclimate reconstructions indicate a warm-wet Medieval Warm period, cold-dry Little Ice age, and relatively humid current warm period around the Chinese Loess Plateau (Fig. 6c, Zhang et al., 2008; Zhao et al., 2010; Tan et al., 2008; Chen et al., 2015). Our data show the long-term evolution of the climate has not significantly increased or reduced the frequency of soil erosion. But this region has been experiencing severe and frequent soil erosion events over the past 2000 BP, whether in enhanced monsoon precipitation periods such as 450-750, 900-1000 and 1950-2000AD or decreased rainfall intervals like 1200-1400, 1650-1800 AD (Fig. 6a, c). Thus, modest climate change has limited influence on soil erosion around the Loess Plateau.

Regional high-intensity rainfall or flood events are more likely to cause soil erosion, since they can carry enough energy to wash a mass of coarse surface soils into the lake centre (Osleger et al., 2009; Parris et al., 2010; Gilli et al., 2013). Around the Tianchi Lake, there are several small gullies on the hillside which can lead the inflow into the lake. When high-intensity rainfall or flood occurred, a large amount of coarse detrital sediments were easily eroded into the lake. By contrast, the suspended fine-grained clay fractions could be washed away from the lake centre (Yu, et al., 2016), or deposited on top of the coarse detrital layers as the hydraulic energy decreased significantly (Gilli et al., 2013). Since the Liupan Mountain is the source region of the downstream rivers, then these heavy rainfall or flood events recorded in Tianchi Lake should also be documented in downstream rivers (Huang et al., 2007, 2011; Zha et al., 2009; Zhang et al., 2014). Thus, a total of 49 flood events in the

downstream rivers were investigated from historical documents over the past 2000 years (Table 2), and the flood frequency per 100yr was also calculated to compare the reconstructed soil erosion records. There are five distinct high-frequency flood periods at 100-300 AD, 500-800 AD, 900-1100 AD, 1200-1500 AD, 1700-1900 AD in the downstream rivers (Fig. 6d). The reconstructed soil erosion coincided well with these flood records in downstream rivers considering dating uncertainties, especially for some individual events such as 200 AD, 300 AD, 600 AD, 750 AD, 900 AD, 1300 AD. Furthermore, the drought/flood (D/F) index in the Longxi area near the Tianchi Lake was chosen to further compare our reconstructed soil erosion records. This index could indicate the local extreme climatic conditions, which was evaluated using a five-level classification method from extreme floods to extreme drought based on local historical documents during the last 2000 years (Tan et al., 2008, 2010). A negative D/F index indicates a higher frequency of flood events, and vice versa. The reconstructed soil erosion records displayed a good relationship with the D/F index. For example, high-frequency floods caused serious soil erosion during 150-300 AD, 450-750 AD, 900-1100 AD, and 1950-2000 AD. Besides, although during the relatively low frequency flood periods at 1200-1400 AD and 1600-1800 AD, these extreme flood peaks also resulted in severe soil erosion events such as 1200 AD, 1300 AD, 1650 AD (Fig. 6e). This evidence supports regional heavy rainfall and floods should be the main cause of soil erosion in the Loess Plateau (Xin et al., 2011; Yu et al., 2016). However, there were also two relatively high frequency flood periods like 1400-1500 AD, 1800-1900AD, Tianchi Lake did not record the

soil erosion events (Fig. 6d, e). Thus, these individual inconsistencies between heavy rainfall/floods and soil erosion suggest there still may be other factors influencing soil erosion in Tianchi Lake.

Reduced vegetation cover could aggravate soil erosion, while increased vegetation cover could stabilize the surface soils (Wang, 1991; Zhou et al., 2008; Chen et al., 2016). Because decreased vegetation cover would make the soils less cohesion and more loose, resulting in more soils being exposed and further eroded in the catchments, vice versa (Fattet et al., 2011; Annabi et al., 2011). A large number of researches showed increasing human activities such as rapidly increased population, expansion of agriculture and excessive land clearing in the highland areas of Liupan Mountain area during the past 2000 years (Ren and Shi, 1986; Ye, 1990; Zhu, 1990; Fang and Zhang, 2007; Zhang et al., 2010). Because of the connectivity between land use, vegetation cover and soil erosion, unsuitable land use and deforestation related to anthropogenic activities can severely aggravated soil erosion through changing any link in connectivity (Rodrigo-Comino et al., 2018; Keesstra et al., 2018b). A high-resolution vegetation evolution record was reconstructed based on fossil pollen from the same sediment core in Tianchi Lake (Zhao et al, 2010; Zhang et al, 2010). The pollen results showed forest began to decrease in the past 2000 years (Fig. 6f, Ren and Shi, 1986; Zhang et al, 2010), indicating sharply reduced vegetation cover. The increased number of archaeological sites during this period further supported vegetation degradation was mainly caused by intensive human distrubances (Zhao et al, 2010). Thus, the tree pollen percentage were chosen to

compare with reconstructed soil erosion records (Fig. 6a, f). Although decreased tree pollen percentage showed correlation with increased reconstructed soil erosion records during 900-1100 AD, 1650-1800 AD, and 1950-2000 AD, it did not display comparable relationship with the reconstructed soil erosion records such as during 150-300 AD, 450-750 AD, and 1300-1400 AD. This might be because that the proxy of tree pollen is expressed by percentage, influenced by the relatively contents of other pollen such as herbs, which is probably more related with the relative variations of vegetation types rather than vegetation cover in the catchment (Zhang et al., 2010; Zhao et al., 2010). Besides, the shift in vegetation types from forest to forest steppe also lags behind deforestation and land clearing caused by human disturbances, thus it is difficult to accurately reflect the vegetation cover at the time.

The concentrations of tree and herb pollen were calculated based on the Lycopodium spore spike to reflect the vegetation cover of trees and herbs around the catchment (Fig. 6g). Four distinct sparse vegetation periods were recorded from both trees and herbs pollen concentrations during the periods of 150-300 AD, 450-750 AD, 900-1400 AD and 1650-2000 AD. Also, the decreased vegetation cover showed good consistency with local dust events inferred from the increasing contribution of EM3, further supporting that vegetation cover does decrease during these periods (Fig. 6g, h). Since decreased vegetation cover would promote surface soil exposure and can easily cause local dust events (Zou and Zhai, 2004; Kimura et al., 2009; Zhou et al., 2015; Liu et al., 2016). Comparison of vegetation cover with soil erosion indicated vegetation deterioration significantly influences soil erosion (Fig. 6a, g). For example,

soil erosion events frequently took place during the light vegetation cover periods of 450-700 AD, 1200-1400 AD, and 1600-1800 AD, in spite of the low frequency of floods (Fig. 6a, d and f). Also, vegetation deterioration further aggravated soil erosion events during the high-frequency flood periods such as 200-300 AD and 900-1000 AD, which could be supported by higher proportion of EM1 indicating stronger soil erosion (Fig. 6a, f). This might be because decreased vegetation cover and loose surface soils, caused by unsustainable land use and deforestation, reduced the capacity of resistance to buffer the high-energy inflow related to heavy rainfall and floods, and lead to high soil erosion (Gyssels and Poesen, 2003; Fattet et al., 2011; Rodrigo-Comino et al., 2018). By contrast, dense vegetation cover could significantly decrease runoff and soil erosion by increasing capacity of resistance and buffering the energy of inflow (Gyssels et al., 2005; Zheng, 2006; Annabi et al., 2011). For examples, the high-frequency floods did not cause serious soil erosion during the periods of the dense vegetation cover at 800-900 AD and 1400-1600 AD. Indeed, a series of simulation experiments supported increasing vegetation covers such forest protection and straw mulch could slow down runoff and then reduce soil erosion (Cerdà et al., 2017; Keesstra et al., 2019). Therefore, our result suggests that regional high-intensity rainfall or flood events are the main cause of soil erosion in the Loess Plateau, whereas human disturbances seriously change land use and influence the vegetation cover, and then aggravate soil erosion.Thus sustainable land management and restoration of catchment vegetation should be encourage to control soil erosion in the Chinese Loess Plateau (Tang, 1991; MEA. 2005; Tsunekawa et al., 2014; Cerdà

et al., 2017; Keesstra et al., 2019).

6. Conclusions

The final four end-members, decomposed from end-member modelling, are deciphered through detailed modern investigations: EM1 indicates catchment soil erosion, EM2 and EM3 likely reflect remote atmospheric dust and local dust fractions, respectively, and EM4 represents the sediment fraction of suspended fine particles that are deposited in slack water conditions. More than 20 strong soil erosion events were inferred from the sediment core and these events could be grouped into six strong soil erosion periods at 150-300 AD, 450-750 AD, 900-1100 AD, 1200-1400 AD, 1600-1800 AD, and 1950-2000 AD. Soil erosion correlates well with intensive flood events in downstream rivers and the neighbouring Longxi area as documented in the historical literature, suggesting that soil erosion is mainly caused by regional high-intensity rainfall or flood events. The good correlation between soil erosion and high content of terrestrial detritus in the sediments also support this. In addition, soil erosion is significantly influenced by the catchment's vegetation cover caused by unsustainable land use and deforestation. Light vegetation cover aggravates soil erosion while dense vegetation cover reduces soil erosion significantly. Our study highlights the restoration of vegetation would be extremely critical to control soil erosion in the Chinese Loess Plateau.

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Figure 1. Location and setting. (a) Map showing the location of our study site Tianchi Lake (TC Lake, red dot) in the south-western Chinese Loess Plateau. Other studies mentioned in the text (dark dots) include Longxi and Wanxiang Cave. The blue lines show the rivers within the boundary of Chinese Loess Plateau (the darks line). (b)

Mean monthly temperature and precipitation from Zhuanglang Station, Gansu, China based on 1971-2000 data from China Meteorological Administration. (c) Photo of Tianchi Lake and sites of the GS07A core (red star) and modern samples: brown dots are lake surface sediments; blue and red dots are surface soils around the lakeshore and hillside, respectively; pink dots are for local dust samples around Liupan Mountain.

Figure 2. Age-depth model and sedimentation rate of core GS07A in Tianchi Lake. Calibrated ages and errors (dark dots with error bars) are given. The modelled age is shown by the red line with the 95% confidence ranges in grey. The red dot near the top of the model is the estimated age from short core GS14A (Zhang, 2016) using ^{210}Pb and ^{137}Cs dating based on grain-size data comparison.

Figure 3. Results of the optimal end-member model. (a) Mean total r^2 across the full size range as a function of the number of end-members to identify the optimal end-member number. (b) Mean total r^2 for each size class for end-member models with 2-6 end members (EM), estimating the stability of each model. (c) Mean total r^2 for each size class showing the goodness-of-fit based on the selected four end-members. (d) Mean total r^2 for each sample showing the goodness-of-fit based on the selected four end-members.

Figure 4. Comparison of grain-size distributions between modelled end-members

with modern samples. (a) Grain-size distributions of modelled end-members with a four end-member model, individual grain-size distribution of each sample from core GS07A, and maximum range for each grain-size class. (b) Grain-size distributions of surface sediments and catchment surface soils. (c) Grain-size distributions of modern dust samples collected around Liupan Mountain. (d) Grain-size distribution of remote atmospheric dust based on data from Sun et al. (2003).

Figure 5. Stratigraphic change of each end-member scores (EM1-EM4), mean grain-size, and the coarse fraction (>100 μm) for core GS07A. The cumulative sum of the four end members is 100%.

Figure 6. Relationship between soil erosion and regional climate change and vegetation cover during the past 2000 years. (a) End-member scores of EM1 from core GS07A. (b) Terrestrial detrital contribution records from core GS07A. (c) Stalagmite $\delta^{18}O$ records (30 point smooth) from Wanxiang cave (Zhang et al., 2008). (d) Flood frequency per 100 yr in downstream rivers including Jinghe River, Weihe River, Fenhe River, middle reaches of the Yellow River, etc (Shi et al., 2002). (e) Drought/flood (D/F) index for the Longxi area (Tan et al., 2008, 2010). (f) Tree pollen from Tianchi Lake (Zhao et al., 2010; Zhang et al., 2010) (g) Concentrations of tree and herbs pollen (Zhang et al., 2010). (h) End-member scores of EM3 (30 point smooth) from core GS07A.

Table 1. AMS-^{14}C dates from Tianchi Lake (core GS07A) in the south-west Chinese Loess Plateau. The dating data are reported by Zhou et al. (2010) and Sun (2011).

Sample No.	Depth (cm)	Material	δ^{13}C (‰VPDB)	^{14}C date ± error (yr BP)	Calibrated age (cal yr BP ± 2σ range)
GS07A1-162	162	leaf	-29.04	680±30	619±109
GS07A1-183	183	leaf	-26.92	855±35	740±91
GS07A1-221	221	leaf	-21.03	1080±35	963±68
GS07A1-260	260	leaf	-11.58	1255±30	1169±81
GS07A2-021	302	leaf	-18.65	1440±45	1378±72
GS07A2-097	382	leaf	-20.96	1775±30	1793±58
GS07A2-148	436	leaf	-24.34	2060±30	2089±81

Table. 2. Historical flood records from the middle reaches of the Yellow River from 1931 AD back to 106 AD, including Jinghe River, Weihe River, Fenhe River, etc (Shi et al., 2002).

Year (AD)	Rivers
1931	Sanhua Reach of Yellow River
1920	Fenhe River, Huihe River
1918	Fenhe River, Danhe River, Sanhua Reach
1917	Zhongtiao Mountain Reach of Yellow River, Qinhe River

1906	Jinghe River, North Luohe River, Weihe River
1898	Southern Shanxi interof Yellow River, Qinhe River, Fenhe River
1888	Weihe River, Sanhua Reach of Yellow River
1887	Helong and Sanhua Reach of Yellow River, Weihe River
1886	Jinghe River, Weihe River, Fenhe River
1832	Sushui River, Sanhua Reach of Yellow River
1819	Helong Reach of Yellow River, Weihe River, Qinhe River, Sushui River
1813	Jinghe River, Weihe River, Fenhe River, Qinhe River
1794	Sanhua Reach of Yellow River, Qinhe River
1775	Fenhe River
1751	Helong Reach of Yellow River, Qinhe River
1745	Weihe River, Sushui River
1737	Fenhe River, Weihe River, Qinhe River
1734	Luohe River, Qinhe River, Sanhua Reach of Yellow River
1730	Helong Reach of Yellow River, Weihe River
1653	Sanhua Reach of Yellow River, Jinghe River, Weihe River
1613	Jinghe River, North Luohe River, Weihe River, Fenhe River, Sanhua Reach of Yellow River
1602	Qinhe River
1570	Helong Reach of Yellow River, Fenhe River
1540	Helong Reach of Yellow River

1508	Helong Reach of Yellow River, Fenhe River
1436	Jinghe River, North Luohe River, Weihe River
1432	Fenhe River
1354	Luohe River, Fenhe River
1345	Luohe River
1340	Luohe River, Sanhua Reach of Yellow River
1099	Weihe River, Sanhua Reach of Yellow River
1076	Sushui River
941	Jinghe River, Helong Reach of Yellow River
939	Sanhua Reach of Yellow River
863	Sanhua Reach of Yellow River
817	Weihe Rive, Fenhe River, Qinhe River
816	Weihe Rive
730	Luohe River, Sanhua Reach of Yellow River
722	Yihe River, Sanhua Reach of Yellow River
720	Sanhua Reach of Yellow River
682	Sanhua Reach of Yellow River, Weihe River
670	Weihe River
655	Sanhua Reach of Yellow River, Weihe River
638	Fenhe River
317	Fenhe River
230	Luohe River, Sanhua Reach of Yellow River

171	Fenhe River, Qinhe River
155	Jinghe River, Weihe River, Sanhua Reach of Yellow River
106	Jinghe River, Weihe River, Sanhua Reach of Yellow River

Sanhua Reach of Yellow River is from Sanmenxia to Huayuankou, Helong Reach of Yellow River is from Hekou Hekou to Longmen area.